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Mesoproterozoic geomagnetic reversal asymmetry in light of new paleomagnetic and geochronological data for the Häme dyke swarm, Finland: Implications for the Nuna supercontinent

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Abstract

Baltica represents one of the key continents of the Mesoproterozoic supercontinent Nuna forming the core of it together with Laurentia and Siberia. This study presents new geochronological and paleomagnetic data obtained for Häme diabase dyke swarm in southern Finland. New U-Pb (baddeleyite) ages 1642 ± 2 Ma and 1647 ± 14 Ma for two reversely magnetized dykes are acquired. Demagnetization revealed a dual polarity remanent magnetization direction carried by magnetite. The combined normal (N) and reversed (R) polarity direction for 11 dykes (=sites) is D = 355.6°, I = -09.1° (k = 8.6 and α95 = 16.6°) yielding a paleomagnetic pole at 23.6°N, 209.8°E (K = 10.6 and A95 = 14.7°) with Van der Voo value Q = 7. N and R magnetized units for the Häme dyke swarm show asymmetry in declination values, probably caused by an age difference between the dykes. The Geocentric Axial Dipole (GAD) model indicates that all geomagnetic reversals should be symmetric (in inclination), yet it has been noted that this is not always the case (e.g. 1.57 Ga Satakunta and Åland dykes in Baltica). By analyzing global dual polarity paleomagnetic data we show that the GAD model is a valid assumption at 1.7 – 1.4 Ga and that the asymmetry between some normal and reversed polarities in global dual-polarity data sets appears randomly over time, and does not follow a global trend. Further, we show that in the case of Åland and Satakunta dykes an unremoved secondary magnetization component could explain the obtained asymmetry. GAD assumption is used to reconstruct the core of Nuna on equatorial latitudes using new data for Häme dykes. Paleomagnetic evidence suggest that maximum assembly of Nuna occurred at 1.5 Ga and the dispersal of the core is proposed to be associated with coeval 1.38 – 1.27 Ga magmatism in its core continents.

Keywords: paleomagnetism, geochronology, Nuna, reversal asymmetry, geocentric axial dipole

1. Introduction

Most cratonic blocks of Earth crust show evidence of collisional events between 2.1 and 1.8 Ga, which has led many researchers to propose that a Mesoproterozoic supercontinent Nuna (a.k.a. Columbia, and Hudsonland; Meert, 2012; Williams et al., 1991, respectively) existed in the Early Proterozoic (e.g. Williams et al., 1991; Hoffman, 1997; Meert, 2002; Rogers and Santosh, 2002; Zhao et al., 2004; Zhang et al., 2012). Many proposed Nuna models differ from each other (e.g., Williams et al., 1991; Hoffman, 1997; Meert, 2002; Rogers and Santosh, 2002; Pesonen et al., 2003; Zhao et al., 2004; Condie, 2004; Zhang et al., 2012; Pisarevsky et al., 2014; Pehrsson et al., in review). However, there is a consensus that Baltica and Laurentia form the core of the Nuna supercontinent in the geologically and paleomagnetically viable North Europe North America (NENA; Gower et al., 1990) connection where northern Norway and Kola Peninsula of Baltica are facing northeastern Greenland of Laurentia between ca. 1.75 and ca. 1.27 Ga (Salminen and Pesonen, 2007; Evans and Pisarevsky, 2008; Lubnina et al., 2010; Pisarevsky and Bylund, 2010; Evans and Mitchell, 2011; Pesonen et al., 2012), but different configurations have also been presented by Johansson (2009) and Halls et al. (2011). Based on Mesoproterozoic passive margins surrounding Siberia (Pisarevsky and Natapov, 2003) and similar geology between Siberia and Western Greenland from 1.9 Ga onward it was recently proposed that Siberia forms the Nuna core together with Baltica and Laurentia in tight fit between East Siberia and Western Greenland (e.g. Rainbird et al., 1998; Wu et al., 2005; Evans and Mitchell, 2011; Ernst et al., 2016; Evans et al., 2016). This tight fit is supported by 1.8 - 1.38 Ga paleomagnetic data from Siberia, Baltica and Laurentia, but an alternative view has also been presented by Pisarevsky et al. (2008). Adding cratons around the core of Nuna has been a major initiative among paleogeographers in recent years (e.g., Congo-São Francisco, Salminen et al, submitted; India, Pisarevsky et al., 2013; North China, Zhang et al., 2012, Xu et al., 2014; Australia cratons, Payne et al., 2009; Li and Evans, 2011; Pehrsson et al., in review; Amazonia, Johansson, 2009; Bispo-Santos et al., 2012; D'Agrella-Filho et al., 2012; 2016) that has lead to the paleogeographical model of Nuna to take shape.

In this study we use the paleomagnetic method, which is the only quantitative tool for a Nuna reconstruction. The number of good quality paleomagnetic data has increased in recent years enabling more reliable global reconstructions based on paleomagnetic data (e.g Evans and Mitchell, 2011; Zhang et al., 2012). Application of paleomagnetic data to the reconstructions requires not only a high reliability of the remanence directions, but also confidence in accurate dating of the rocks and their acquisition of remanent magnetization. A reliable paleomagnetic pole generally fulfills at least three of the seven quality criteria of Van der Voo (1990). If two of these include an adequately precise geochronological age and a positive paleomagnetic field test, the obtained paleomagnetic pole can be called a "key" pole (Buchan et al., 2000; Buchan, 2013). Paleomagnetic data for different continents allow comparison of lengths and shapes of apparent polar wander paths (APWPs) to test proposed long-lived proximities of the cratons. As long as these landmasses have traveled together they should have identical APWPs. In the absence of well-defined APWPs, pairs of coeval paleomagnetic poles from these cratons can be used for a rough test (e.g. Buchan et al., 2000; Evans and Pisarevsky, 2008). These pairs of paleopoles should plot on top of each other within their error limits, after Euler rotation to the continents' correct relative configuration.

To further test the NENA connection, the Early Mesoproterozoic 1.65 Ga Häme dykes that are related to rapakivi granite magmatism in southern Finland are studied, since they can provide a good quality coeval paleopole to the 1.63 Ga Melville Bugt pole (Halls et al., 2011) for Laurentia. A paleomagnetic study on the Häme dyke swarm has been carried out earlier, however only a secondary component was obtained (Neuvonen, 1967) most probably due to the inadequate demagnetization method that was used (single-step AF demagnetization at 30 mT). Here we use modern demagnetization techniques combined with U-Pb isotope geochronology and extend the analysis to studying the geomagnetic polarity asymmetry, using early Mesoproterozoic paleomagnetic data.

The axial dipolar field model of Earth's geomagnetic field is the key assumption for interpreting the past latitude and geography of the continents from paleomagnetic data. The model indicates that all geomagnetic reversals should be symmetric, meaning that obtained normal (N) and reversed (R) magnetization directions of dual polarity data are antiparallel. Notable asymmetry has been obtained earlier at 1.1 Ga in the Keweenawan rocks of Laurentia (e.g. Palmer, 1970; Pesonen and Nevanlinna, 1981; Halls and Pesonen, 1982; Pesonen and Halls, 1983; Nevanlinna and Pesonen, 1983; Schmidt and Williams, 2003), but Swanson-Hysell et al. (2009) showed that it was an artefact of the rapid motion of North America during this time. In addition, pronounced asymmetric paleomagnetic results have recently been obtained for Mesoproterozoic diabase dykes in Åland and Satakunta, southern Finland (Salminen et al., 2014; 2015). Several different reasons could explain this: i) unusual behavior of the geomagnetic field, especially permanent non-dipolar field contamination (e.g., Pesonen and Neuvonen 1981;Veikkolainen et al., 2014a,b), ii) an unremoved secondary component (Halls et al., 2011); iii) an age difference (associated with tectonic drift) between normal and reversed polarity directions (Swanson-Hysell et al., 2009), iv) relative crustal tilting of blocks with N- and R- polarity directions (Halls and Shaw, 1988). A further aim of this study is to test the possibility of a non-dipolar field contamination during the Mesoproterozoic and to analyze the effect of secondary components to dual polarity data.

2. Geological background

The Häme dyke swarm is located in the Svecofennian domain of the Fennoscandian Shield in southern Finland. The Svecofennian domain was formed at 1920-1870 Ma as a result of accretion of several island-arcs and microcontinents against the Archean craton in the NE (Lahtinen et al., 2005). Lithologically the rocks are predominantly composed of Paleoproterozoic metasedimentary and metavolcanic rocks of island-arc type, and granitoids which have intruded into these supracrustal rocks. The sedimentary and volcanic rocks were metamorphosed under high temperature and low-pressure conditions (Korsman, 1977; Korsman et al., 1984). After the stabilization of the Svecofennian domain, about 200 Ma later, the crust in southern Finland was intruded by Mesoproterozoic rapakivi granite batholiths and stocks that sharply cut the surrounding Paleoproterozoic bedrock (e.g. Rämö 1991). The total age range from U-Pb dating for the Fennoscandian rapakivi province is ca. 1.65 – 1.5 Ga. The rapakivi granites are associated with diabase and quartz porphyry dyke swarms that radiate from the rapakivi granite (Fig. 1). We will use the term “Subjotnian” later in this paper when we refer to these Early Mesoproterozoic units in the southern part of Finland. The Häme diabase dyke swarm is associated with the Ahvenisto rapakivi granite which forms a separate satellite complex north of the colossal Wiborg rapakivi granite batholith. The Ahvenisto satellite consists of a horseshoe-shaped gabbro-anorthosite rim and a central rapakivi granite batholith in which the major rock type is biotite granite (Savolahti, 1956). The rapakivi granites in southeastern Finland were emplaced at shallow crustal levels over long periods. The Wiborg, Suomenniemi and Ahvenisto granites have U-Pb zircon ages of 1650-1620 Ma (e.g. Vaasjoki et al., 1991) and according to Rämö et al. (2014) the emplacement of the plutonic rocks of the Wiborg batholith took place 12 Ma at minimum.

Most diabase dykes around the rapakivi areas of southeastern Finland are considered to be coeval or slightly older than the rapakivi granite, but genetically associated. The contemporaneous diabase dykes presumably represent derivatives of the mantle-originated thermal perturbations that caused anatexis of deep parts of the crust and subsequent emplacement of rapakivi granite batholiths in an extensional tectonic setting (Rämö, 1991; Haapala and Rämö, 1992). The rapakivi granites are shown to cut the diabase dykes. Presumably the upward movement of rapakivi granite melt continued after the injection of the diabase dyke magma and eventually the rapakivi massifs cut the diabase dykes (Laitakari, 1969). According to Laitakari and Leino (1989) the Ahvenisto gabbro anorthosite pluton probably formed the magma chamber for two types of diabase dykes (see below) radiating from the Ahvenisto pluton. The intrusion of rapakivi massifs disrupted the bedrock around their margins by forming steep faults and joints that were filled with basaltic magma. However, the diabase magma most likely intruded along structures that best agreed with the direction of the deep faults, and the jointing systems of the bedrock may thus be older than the diabase dykes (Laitakari, 1969)

The Häme dyke swarm extends ca. 150 km NW from the Wiborg and Ahvenisto rapakivi granites. Several earlier geochronological results on the dyke swarms related to Wiborg rapakivi have been reported (Vorma, 1975; Vaasjoki, 1977; Laitakari, 1987; Siivola, 1987; Vaasjoki and Sakko, 1989; Vaasjoki et al., 1991; Heinonen et al., 2010). The most reliable ages are the 1646 ± 6 Ma (U-Pb, zircon) for the Ansio diabase dyke trending N60W (Häme swarm; Laitakari, 1987); the 1635 ± 3 Ma (U-Pb, zircon) for the Nikkari quartz porphyritic dyke (Suomenniemi swarm; Vaasjoki et al., 1991); 1643 ± 5 Ma

(U-Pb, zircon) for the Lovasjärvi diabase dyke (Suomenniemi swarm; Siivola, 1987); and 1636 ± 2 Ma (U-Pb, zircon) for the Leviänlahdenvuoret quartz porphyritic dyke (Ahvenisto swarm; Heinonen et al., 2010). These are coeval with a gabbro-anorthosite surrounding the Mäntyharju rapakivi intrusion (Vaasjoki and Sakko, 1989). Older U-Pb zircon ages of 1690 Ma for the Hyvärilä diabase dyke (Ahvenisto swarm; Vorma, 1975), and 1667 ± 9 Ma for the N80°NW Virmaila dyke (Häme swarm; Vaasjoki et al., 1991) were obtained. However they are considered unrealiable in light of the age determinations shown in this study.

The widths of the Häme dykes vary from 250 m to a few centimetres (Laitakari, 1969). The dykes comprise two dyke sets with different trends and compositions (Laitakari, 1969, 1987; Luttinen and Kosunen, 2006, Vaasjoki and Sakko, 1989). One dyke set is about 100 km long and have dominating strikes of 45°- 60°NW. The other dykes trend in 85°NW direction and can be followed for about 150 km from the southern part of the Ahvenisto pluton through Orivesi up to Kuru. According to Laitakari (1969) there are several observations where the trend is between these maxima (60 - 85°NW). The dykes dip vertically or subvertically. In general, the widths of the dykes get narrower the longer the distance is from the rapakivi granite. Contacts with the host rock are typically sharp. Non-crystalline glass has been discovered in some of the narrowest dykes (Lindqvist and Laitakari, 1980). In some of the dykes there are amygdoloids which indicate crystallization of magma close to the surface (Laitakari, 1987).

Compositionally the Häme diabase dykes are unmetamorphosed olivine diabases where the main minerals are plagioclase, olivine and clinopyroxene (Laitakari, 1987). The dykes may also contain biotite, potassium feldspar and orthopyroxene. The 80°NW trending diabase dykes are mainly olivine tholeiites and typically contain abundant olivine, but no plagioclase phenocrysts or big plagioclase fragments (Laitakari and Leino, 1989). The 60° NW trending dykes are characterized by phenocrysts, megacrysts (up to < 20 cm) and fragments of plagioclase (Laitakari, 1969) which due to flowage differentiation are concentrated in the central parts of the dyke.

3. Sampling and Methods

3.1 Sampling

Standard 2.5-cm diameter cores were collected with a portable field drill from 22 diabase dykes of Häme swarm for paleomagnetic measurements during field campaigns in 2009 and 2014 (Fig. 1). Host rocks were sampled for baked contact tests (Everitt and Clegg, 1962) at 11 of the dyke sites. Additional unbaked host rock sites were sampled in 2016. Cored samples were oriented using solar and/or magnetic compasses. One new geochronology sample was taken from the more than 60 m wide Torittu dyke (site H17) for U-Pb dating. The Torittu dyke has ophitic texture which is spotted due to plagioclase clusters, 1-2 cm in diameter (Laitakari, 1969). The interstices between the plagioclase laths are filled by olivine and augite. Accessory minerals are titanomagnetite, biotite, apatite, zircon and serpentine as the alteration product of olivine. Another new age dating was done on existing baddeleyite from Virmaila dyke (VR, width 60 m) that was reanalyzed using Isotope Dilution Thermal Ionization Mass Spectrometry (ID-TIMS) method.

3.2 Geochronological methods

A ca. 5 kg whole-rock sample from a Torittu dyke (H17) were prepared for dating. Zircon and baddeleyite for Laser Ablation Multicollector Inductively Coupled Plasma Mass Spectrometry (LA-MC-ICPMS) U–Pb dating were selected by hand-picking after heavy liquid (CH_2I_2, and Clerici solution) and Frantz magnetic separation. The dyke sample yielded 20 baddeleyite grains (width > 20 μm) and 15 zircon grains. The chosen grains were mounted in epoxy resin and sectioned approximately in half and polished. Back-scattered electron images (BSE) and cathodo luminescence (CL) images were taken using SEM (Scanning Electron Microscope) to target the spot analysis sites on the mineral grains. This sample was assigned sample code A2414 for the Finnish Rock Age Database of the Geological Survey of Finland (GTK).

U–Pb dating analyses for Torittu sample were performed using a Nu Plasma HR multicollector ICPMS at the Geological Survey of Finland in Espoo using a technique very similar to Rosa et al. (2009) except that an Analyte G2 193 nm laser laser microprobe was used. The analyses were made in static ablation mode with a beam diameter of 20 µm, pulse frequency of 10 Hz, and beam energy density of 2.07 J/cm^2. A single U–Pb measurement included 30 s of on-mass background measurement, followed by 60 s of ablation with a stationary beam. Masses 204, 206 and 207 were measured in secondary electron multipliers and 238 in the extra high mass Faraday collector. Ion counts were converted and reported as volts by the Nu Plasma time-resolved analysis software. ^{235}U was calculated from the signal at mass 238 using a natural ^{238}U/^{235}U=137.88. Mass number 204 was used as a monitor for common ^{204}Pb. Raw data were corrected for background, laser induced elemental fractionation, mass discrimination, and drift in ion counter gains and reduced to U–Pb isotope ratios by calibration to concordant reference baddeleyite of known age, using protocols adapted from Andersen et al. (2004) and Jackson et al. (2004). In-house standard baddeleyite A974 (1256.2 ± 1.4 Ma; Söderlund et al., 2004) was used for calibration. Age related common lead (Stacey and Kramers, 1975) correction was used when the analysis showed common lead contents above the detection limit. The calculations were done off-line, using an interactive spreadsheet program written in Microsoft Excel / VBA by Tom Andersen (Rosa et al., 2009). To compensate for drift in instrument sensitivity and Faraday vs. electron multiplier gain during an analytical session, a correlation of signal vs. time was assumed for the reference zircons. A description of the algorithms used is provided in Rosa et al (2009). The U−Pb isotopic data were potted and the age calculations were performed using the Isoplot/Ex 3 program (Ludwig, 2003). Ages were calculated with 2σ errors and without decay constants errors. Data-point error ellipses in the Fig. 2 are at the 2σ level.

Another new age dating was done on existing baddeleyite from Virmaila dyke (VR, width 60 m) that was reanalyzed using Isotope Dilution Thermal Ionization Mass Spectrometry (ID-TIMS) method. Earlier analyses for Virmaila diabase dyke had yielded an age of 1667 ± 9 Ma (Vaasjoki and Sakko, 1988). We reanalyzed the existing baddeleyite from Virmaila dyke (VR, width 60 m). The decomposition of baddeleyite and extraction of uranium and lead for Isotope Dilution Thermal Ionization Mass Spectrometry (ID-TIMS) age determinations mainly follows the procedure described by Krogh (1973, 1982). ^{235}U-^{208}Pb-spiked and unspiked isotopic ratios were measured using a VG Sector 54 TIMS. The measured lead and uranium isotopic ratios were normalized to the accepted ratios of SRM 981 and U500 standards. The Pb/U ratios were calculated using the PbDat program (vers.1.24; Ludwig, 1993). The concordia plots and the final age calculations were done using the Isoplot/Ex 3.00 program (Ludwig, 2003). The common lead corrections were done using the age related Stacey and Kramers (1975) lead isotope compositions ($^{206}Pb/^{204}Pb$±0.2, $^{208}Pb/^{204}Pb$±0.2, and $^{207}Pb/^{204}Pb$±0.1). The total procedural blank level was 20-50 pg. All the ages are calculated with 2σ errors and without decay constant errors. In Fig. 3, the data-point error are at 2σ level.

3.3 Rock magnetic and paleomagnetic methods

Rock magnetic and paleomagnetic measurements were carried out at the Solid Earth Geophysics Laboratory of the University of Helsinki (UH), Finland. Magnetic mineralogy was investigated by thermomagnetic analysis of selected powdered whole-rock samples.

Temperature dependence of low-field magnetic susceptibility was measured from -192°C to ~700°C (in argon gas) followed by cooling back to room temperature using an Agico CS3-KLY-3S Kappabridge system, which measures the bulk susceptibility of the samples. Curie temperatures were determined using the Cureval 8.0 program (http://www.agico.com).

Stepwise alternating field (AF) demagnetizations were done using a three-axis demagnetizer with maximum field up to 160 mT, coupled with a cryogenic 2G (now WSGI) DC SQUID magnetometer to isolate the characteristic remanent magnetization (ChRM) component. Sister specimens were thermally demagnetized using an argon-atmosphere ASC Scientific model TD-48SC furnace. Remanent magnetizations were measured with a DC SQUID magnetometer. Vector components were visually identified using stereographic and orthogonal projections (Zijderveld, 1967) and the directions were calculated by a least squares method (Kirschvink, 1980). Mean remanence directions for the different components were calculated according to Fisher (1953), giving a unit weight to each sample (each specimen has a unit weight within a sample) to compute site mean directions and corresponding virtual geomagnetic poles (Irving, 1964). The paleogeographical reconstructions and apparent polar wander paths (APWP) were done with the GPlates program (Boyden et al., 2011; Gurnis et al., 2011; Williams et al., 2012).

4. RESULTS

4.1 U-Pb geochronology results

Results for LA-MC-ICPMS U-Pb baddelyite data from the Torittu diabase dyke sample is presented in Table 1 and in Fig. 2. Only one small elongated magmatic zircon grain was dated, yielding an age of 1.90 Ga. This is either an inherited grain or contamination from the rock crushing and separating processes, and will not be discussed further. Eleven baddeleyite domains were dated, all of which were euhedral and translucent. Two analyses showing major discordance and high common lead contents (2.6% and 3.5%, respectively) were rejected (Table 1). We further ignored the two "youngest" data points with lower error correlation values and moderate common lead contents (1.9% and 2.2%, respectively) and calculated a concordia age of 1647 ± 14 Ma (MSWD=0.03; n=7) (Fig. 2).

Results in the reanalysis of the baddeleyite from the Virmaila dyke two fractions of best existing baddeleyite of diabase sample of Virmaila dyke using the ID-TIMS is presented in Table 2 and in Fig. 3. Two fractions of the best quality baddeleyite grains were air-abraded for 30 minutes. The U-Pb TIMS measurements (Table 2) yielded concordant and nearly concordant ages with a mean of 1642 ± 2 Ma (Fig. 3) being notably younger than the earlier age of 1667 ± 9 Ma (Vaasjoki and Sakko, 1989).

4.2 Rock magnetic results

Examples of representative rock magnetic results are shown in Fig. 4. Temperature dependence of low-field magnetic susceptibility at low temperatures (heating from -192°C to room temperature) shows the presence of nearly stoichiometric magnetite in diabase samples indicated by the Verwey transition (Verwey, 1939) at around -153°C to -146°C. Temperature dependence of low-field magnetic susceptibility at high temperatures (in argon) show the presence of Hopkinson peak and Curie temperatures at range of 570 - 582°C being consistent with magnetite and/or low-Ti titanomagnetite as the magnetic carrier. Small irreversibility between heating and cooling curves may be due to slight mineralogical changes during the heating.

4.3 Paleomagnetic results

4.3.1 Primary remanent magnetization

Paleomagnetic results are listed in Table 3, and representative demagnetization behaviors are illustrated in Figs 5 and 6. We measured a total of 125 diabase samples from 22 dykes; total of 44 baked host rock samples for 11 of the dyke sites and 33 unbaked samples. Eleven dykes (45 samples) show presumably primary characteristic remanent magnetization (ChRM) dual-polarity direction with northerly and southerly declinations and shallow up- and downward pointing inclinations (Fig. 7). The remaining sampled Häme dykes were remagnetized or showed unstable directions (see Table 3). In general, samples showing primary ChRM were stable to both AF and thermal demagnetization methods. ChRM was obtained with 30-160 mT AF fields and with unblocking temperatures up to 585°C. We accepted the data from dykes where at least two samples show similar directions. Ideally components with Mean Angular Deviation (MAD) values less than 5° were included, but in some cases a higher MAD was accepted if a clear ChRM was observed. The mean direction for ChRM of five normal polarity Häme dykes is D = 015.8°, I = -09.2° (with k = 10.8, α_{95} = 24.3°) and the palaeomagnetic pole derived from VGPs is at 22.4°N, 187.9°E (with K = 14.8 and A95 = 20.6°). For six reversed polarity Häme dykes the ChRM mean is D = 159.3°, I = 8.1° (with k = 17.0, α_{95} = 16.7°) (Table 3, Fig. 7) and the palaeomagnetic pole is at 22.3°N, 227.4°E with K = 33.5 and A95 = 11.7°. Two of the dated dykes, Virmaila (1642 ± 2 Ma) and Torittu (1647 ± 14 Ma), show reversed polarity directions. The combined grand mean normal and reversed polarity ChRM direction for 11 sites is D = 355.6°, I = -09.1° (with k = 8.6 and α_{95} = 16.6°) yielding a paleomagnetic pole at 23.6°N, 209.8°E (with K = 10.6 and A95 = 14.7°).

Since the primary remanent magnetization of the dykes has thermal origin by default, the baked contact test (Everitt and Clegg 1962) can be used to verify the primary nature of magnetization. Baked-contact tests were performed at eleven dyke sites of the Häme region, but due to unstable remanent magnetization in the host rocks it was successful only for the dated 1642 ± 2 Ma Virmaila dyke. Reversely magnetized Virmaila dyke (sites VR, HR, H8) provides a full positive baked contact tests (Fig. 6, Table 3). The dyke and the baked host rock (Figs 1 and 6, Table 3) show a shallow southerly (reversed polarity) pointing ChRM directions. The unbaked rocks at the sampling area were not the best targets for paleomagnetic study, but we were able to obtain a typical Svecofennian type direction (D=341°, I=49°) from a Svecofennian aged host granodiorite, 2.5 km and ca. 1.0 km from the Virmaila dyke (Fig. 6, Table 3).

4.3.2 Secondary remanent magnetization

In addition to the primary ChRM most of the measured samples show secondary magnetization components. In many cases a viscous component resembling the Present Earth's Field (PEF) direction at the sampling site (D = 7°, I = 73°) was removed with fields ≤10 mT and temperatures <200°C. The majority of the samples also show another secondary component, here after called component B following Mertanen (1995), with an easterly declination and an intermediate to steep downward inclination (Fig. 7). In some cases component B was obtained from the coarser grained samples taken from the interior of the dyke and it was removed with a field of ≤20 mT and temperatures <350°C indicating its secondary nature. In these cases a shallow primary component was obtained for samples from the finer grained margins of dyke. In other cases higher AF field (30 -160 mT) and higher temperatures 500-580°C were needed to demagnetize component B and an original remanent magnetization direction was not obtained at all in the dyke (e.g. dyke H11). Due to the positive baked contact test the dual-polarity shallow component was interpreted to represent a primary magnetization and thereafter the component B is interpreted to represent a secondary magnetization, being similar to the one obtained in several other intrusions and shear zones in Fennoscandia (e.g. Mertanen, 1995; Preeden et al., 2009; Salminen et al. 2014; 2015).

Three dykes (H16, H19 and H22) show high coercivity remanent magnetization directions with southerly declinations and steep downward inclinations (Fig. 7) and two dykes (H3 and H7) show high coercivity WNW declinations with downward shallow inclinations. The origins of these directions are not known and they are not further discussed here.

5. DISCUSSION AND CONCLUSIONS

5.1 Geochronology

Sampled Häme dykes are associated with Wiborg rapakivi within the Fennoscandian Shield. Age succession of Wiborg rapakivi granite from oldest in the North (1646 ± 4 Ma Värtö tirilite) and youngest in the South (1627 ± 3 Ma Ristisaari dark wiborgite) suggest that the overall locus of magmatic activity may have shifted southward during the build-up of the Wiborg batholith (Rämö et al., 2014). According to Laitakari (1969) and Vaasjoki and Sakko (1989) two sets of Häme diabase dykes occur, trending either 60°NW or 80°NW. Dykes trending 60°NW are more differentiated and they often contain megacrysts of plagioclase. The N80W striking diabase set is more uniform, consisting mainly of medium grained ophitic rock.

This study provides a new U-Pb (baddeleyte) age of 1647 ± 14 Ma for a diabase dyke in 80°NW Torittu (Häme swarm) and reanalysing the existing sample from Virmaila dyke provides a new age of 1642 ± 2 Ma being clearly younger than the previous U-Pb (baddeleyite) age of 1667 ± 9 Ma (Vaasjoki and Sakko, 1988) for Virmaila. One reason for this discrepancy could be the baddeleyite fraction might have included some inherited zircons. The new age narrows down the emplacement time of Häme dyke swarm with slightly different trend of strikes and now the petrologically different dyke sets show coeval ages (Vaasjoki and Sakko, 1989). See as well Rämö et al. (2014) and Rämö and Mänttäri (2015).

5.2 Häme paleomagnetic data

5.2.1 Quality of the Häme paleomagnetic pole

Earlier only magnetization direction close to present Earth's field (PEF) direction has been obtained for Häme dykes (Neuvonen, 1967). Since the 1960s demagnetization techniques have developed, and with more extensive sampling, a primary magnetization component has now been obtained for the newly dated Häme dyke swarm (1642 ± 2 Ma, 1647 ± 14 Ma U-Pb on baddeleyite). The new pole for the Häme swarm lies at 23.6°N, 209.8°E (with K = 10.6 and A95 = 14.7°). This new pole fulfils seven of the Van der Voo (1990) reliability criteria providing a new key pole (e.g. Buchan and Halls, 1990; Buchan et al., 2000; Buchan, 2013) for Baltica. This stresses the limits for the required statistics, and for the baked contact test the Svecofennian aged (1.9-1.8 Ga) host rock in SE Finland is a poor carrier of remanence. The pole has: (1) well-determined U-Pb (baddeleyite) ages of 1642 ± 2 Ma and 1647 ± 14 Ma for reversely magnetized Virmaila and Torittu dykes, respectively. This is also interpreted to represent the age of magnetization; (2) The statistics of this new combined R- and N - polarity pole is just within the limits of Van der Voo (1990) criteria #2 (more than 24 samples, K>10, and A95 <16°), but fulfils them (45 samples, K=10.6, and A95 =14.7°). The statistics of the reversed polarity pole are better than the statistics for normal polarity pole. The reversed polarity pole fulfils criteria #2 (25 samples, K=33.5, and A95 =11.7°), but the normal polarity pole does not (20 samples, K=14.8, and A95 =20.6°); (3) Adequate AF and thermal demagnetization methods were used; (4) Reversely magnetized Virmaila dyke demonstrates a full positive baked contact test. Subjotnian R-polarity directions for Virmaila dyke itself are scattered with α_{95} = 38.2°, but they are clearly Subjotnian. Since there are only four specimens taken from the rather coarse grained part of this 60 m wide dyke showing Subjotnian direction, the directions can be more scattered and α_{95} larger than when more than four specimens show a similar direction. The baked host rock samples (seven samples) with

α_{95} = 18.3° show better statistics than the dyke itself. For unbaked host rock samples the quality of data is not as good as for baked host samples, because the majority of the samples show rather unstable behaviour during demagnetization. It has been noted earlier that the Svecofennian aged (1.9-1.8 Ga) host rock in SE Finland is a poor carrier of remanence (e.g. Mertanen, 1995). Although only two unbaked host rock samples for the reversely magnetized Virmaila dyke show Svecofennian direction, their existence clearly demonstrates that the host rock carries different remanence than the dyke and that this remanence has a Svecofennian origin. Moreover, the Svecofennian component was also obtained for an unbaked host rock sample for the N polarity H1 dyke (Table 3). No stable direction for the baked host rock for H1 dyke was obtained, but the unbaked host rock sample demonstrates clearly the presence of a Svecofennian aged component. Despite the scatter of data the full positive baked contact test is demonstrated; (5) Structural control; and (6) the presence of reversals, although they do not pass the reversal test (McFadden and McElhinny, 1990). There are several case of primary remanences from a single magmatic event where reversals are not antipodal because emplacement extends over a significant time interval during which the continent was in motion (e.g. Buchan, 2013). For example, the 2.13-2.10 Ga Marathon dyke swarm of North America was emplaced over 25 Ma and shows polarity asymmetry (Halls et al., 2008). Our interpretation is that to give Q6 = 1 (presence of reversals) does not require the presence of antipodal reversals. The criteria number (7) - a similarity to younger poles is satisfied as well. The pole is similar to the poles obtained from Carboniferous and Permian formations in Sweden and Norway (e.g. Torsvik et al., 1992). However, these areas are more than 500 km to the southwest of our sampling sites, and considering the positive baked contact tests for the dykes it is very unlikely that Carboniferous-Permian geological events generated the ChRM in the Häme dykes.

5.3 Asymmetry in Mesoproterozoic paleomagnetic record and geomagnetic field properties

Site mean remanent magnetization values for Häme dyke intrusions show asymmetry, i.e. the mean normal (N) and reversed (R) directions are not antiparallel at 95% confidence level in declination values (Fig. 7). Mean declination for normal polarity dykes is D_N = 009.0° and for reversed polarity dykes declination is D_R = 159.3° (339.3° when polarity is reversed to normal polarity). Inclination agrees for both polarities. Non-antiparallel directions have already been obtained earlier in other Subjotnian studies of Fennoscandia (e.g. Pesonen and Neuvonen, 1981; Salminen et al., 2014; 2015), but the asymmetry was shown in inclination values as in case for Åland and Satakunta dykes (Fig. 8). A classical example of asymmetry is the 1.1 Ga Keweenawan rocks in North America (e.g. Palmer, 1970; Halls and Pesonen, 1982; Pesonen and Halls, 1983; Schmidt and Williams, 2003), which was first interpreted to be due to anomalous behavior of the geomagnetic field (e.g. Palmer, 1970; Halls and Pesonen, 1982; Pesonen and Halls, 1983; Schmidt and Williams, 2003), but was later shown to be a result of the rapid motion of North America during this time (Swanson-Hysell et al., 2009). We further studied the asymmetry in the Häme and other Subjotnian dyke swarms in Finland, and discuss possible reasons that could explain the data. First, the effect of secondary magnetization to distort the remanence directions was studied.

5.3.1 Secondary magnetizations

Some of the Häme dykes show secondary magnetizations, either a viscous component in the present Earth's field (PEF) direction or the NE directed intermediate inclination component B, which is common in several units in Fennoscandia (e.g. Mertanen, 1995; Salminen et al., 2014; 2015). We tested the effect of an unremoved secondary component through vector analysis, by subtracting the measured secondary component vector from the obtained normal